LIMITED CIRCULATION

This paper is not an official document; it is for information only and must not be quoted in any printed publication.

AGRICULTURAL MEMORANDUM NO. 990

# Modelling the flow of moisture in a soil

## Part 1 - Basic concepts

by M Field

### 1. Soil Water Potential

Water moves in the soil if there is a gradient of potential energy - the water moves from a point of higher potential to one of lower potential. The water potential at a point can be written as,

$$\bar{\Psi} = \psi + \psi_g \qquad 1.1$$

where $\psi_g$ is the gravitational potential and $\psi$ is the matric or pressure potential.

$\psi_g$ is directly proportional to the height of the point above a given reference datum.

If $\psi$ is negative it is known as the matric potential. $\psi$ is negative in unsaturated soils, e.g. above a water table, because energy is needed to remove water. If $\psi$ is positive it is known as the pressure potential. $\psi$ is positive in saturated soils, e.g. below a water table.

Water Potentials are often expressed not as a true energy but as an 'energy per unit weight of soil water'. The units then become more conveniently those of distance. Under this scheme water at a pressure of 100 $Nm^{-2}$ or 1 mbar has an 'energy potential' of 1 cm (taking $g = 10\ ms^{-1}$ ).

If we measure distances positively in an upward direction then $\psi_g = z$ cm.

We can then write,

$$\bar{\Psi} = \psi + z \text{ cm} \qquad 1.2$$

where $\psi$ is called the soil moisture pressure head $P$ if $\psi$ is positive or the suction $S = -\psi$ if $\psi$ is negative. $\bar{\Psi}$ now has the units of distance.

To measure pressure head $P$ in the soil a small bore tube known as a piezometer can be used. To measure suction $S$ a tensiometer can be used.

The way these simple instruments function serves to illustrate the concepts outlined above (12).

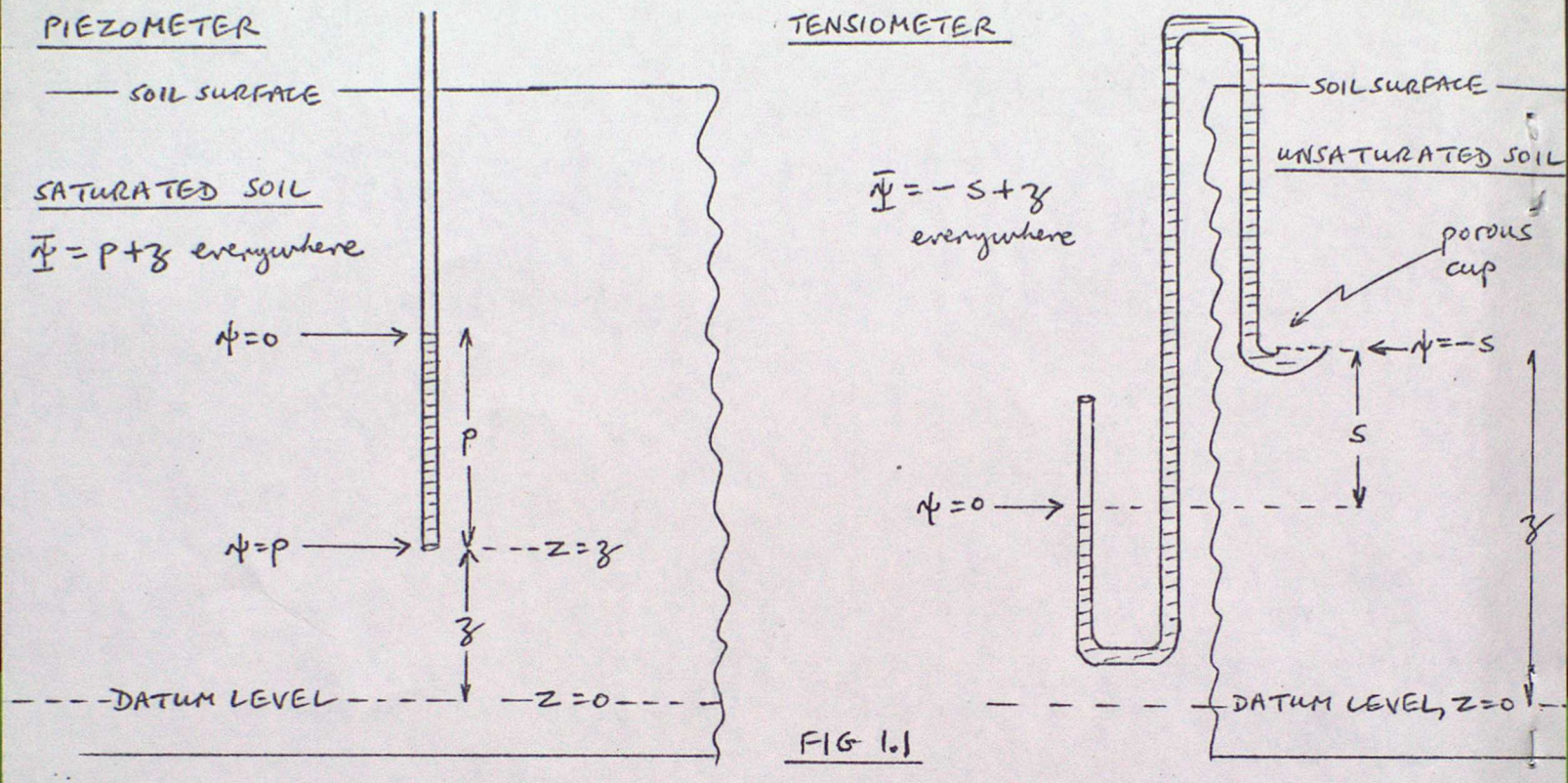


FIG 1.1

## 2. Soil Moisture Characteristic

At zero pressure head all the pores in the soil are filled with water. As the suction increases from zero the pores begin to empty starting with the largest ones and the Moisture Content falls from its saturation value. The relationship between the Suction and the Moisture Content is known as the Soil Moisture Characteristic.

The Moisture Content of a sample of soil is often expressed as $\theta_v$ where,

$$\theta_v = \frac{\text{volume of water in soil sample}}{\text{volume of soil sample}} \qquad 2.1$$

In general the Soil Moisture Characteristic for a drying soil (desorption curve) is not the same as for a wetting soil (adsorption curve) due to hysteresis effects. However often the changes to the Soil Moisture characteristic due to hysteresis are smaller than those due to soil variability.

The range of suctions experienced by a soil varies from 0 to about $10^7$ cm.

A logarithmic scale is sometimes used to express this making use of the quantity pF defined by

$$pF = \log_{10} (S) \qquad 2.2$$

Permanent Wilting Point (PWP) is defined as a suction of 15 bar or 15000 cm.

Field Capacity (FC) is sometimes considered to be a suction of 50 cm, although this is a poor definition (see later).

The Available Water is $AW = \theta_v(FC) - \theta_v(PWP)$ 2.3

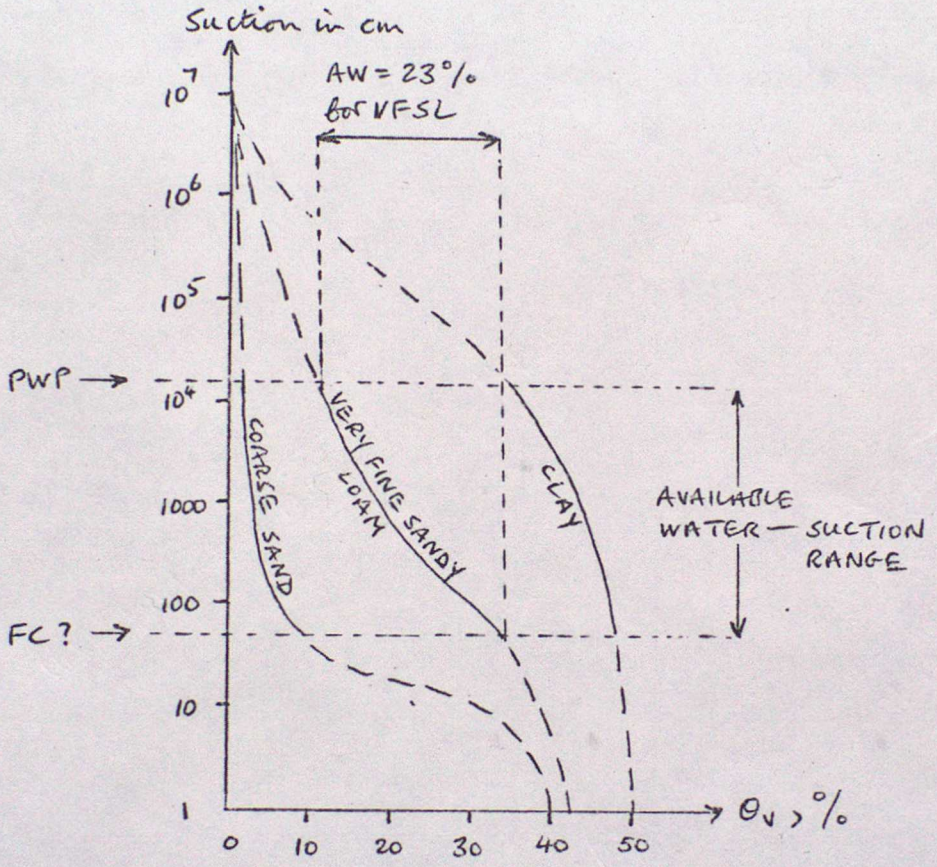


FIG 2.1 SOIL MOISTURE CHARACTERISTICS

Consider a coarse sand and a clay soil which have reached equilibrium with their respective water tables at a depth of 100 cm. The suction at the surface is 100 cm. At this level $\theta_v = 0.06$ for the sand and $\theta_v = 0.48$ for the clay. For a water table in equilibrium at 50 cm, the respective Moisture Contents at the surface for sand and clay are 0.10 and 0.49.

## 3. Hydraulic Conductivity and D'arcy's Law

D'arcy's law relates the volumetric flux of water to the gradient of Soil Water Potential $\bar{\Psi}$ ,

$$v = -k\,\partial/\partial z(\bar{\Psi}) \qquad 3.1$$

where $V$ = volumetric flux of water (e.g. $cm^3$ $cm^{-2}$ $day^{-1}$)
and $k$ = hydraulic conductivity, expressed as a function of $S$ or $\theta_v$.

D'arcy's law can be used to define $k$.

Substituting eqn. 1.2 we find simply that,

$$V = -k\left(\frac{\partial\psi}{\partial z}+1\right) \qquad 3.2$$

For a soil in equilibrium with a water table, $\frac{\partial\psi}{\partial z} = -1$ in the unsaturated zone and hence $V = 0$ as required.

An example of 3 Hydraulic Conductivity curves appears in fig 3.1

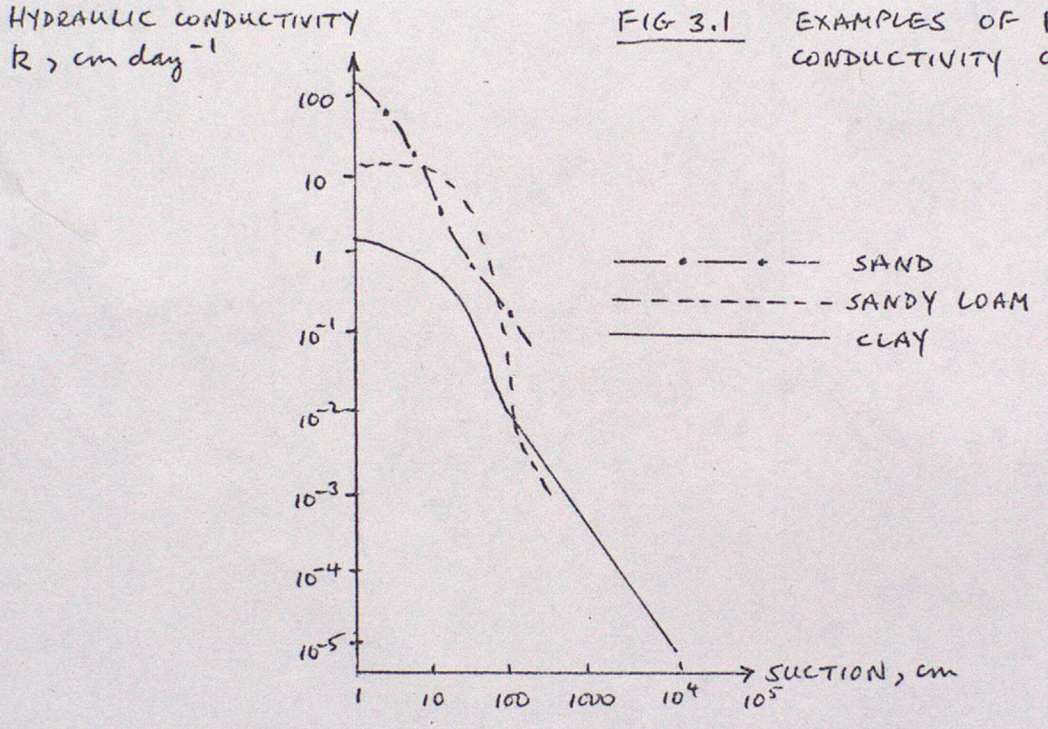


FIG 3.1 EXAMPLES OF HYDRAULIC CONDUCTIVITY CURVES

## 4. Field Capacity and Soil Moisture Deficit

A Soil Moisture Deficit is generally thought of as the amount of water that a soil can accept without incurring drainage losses from the base of the profile. Several criteria are implicit in this concept,

i. There must be a reproducable Moisture Content for the profile with respect to which the SMD is quantified - the Field Capacity datum.

ii. The Hydraulic Conductivity of the soil must be sufficiently great to enable any surfeit of water to drain from the profile within a reasonably short period, normally taken as 48 hours.

iii. The depth of the profile under consideration must be defined. This can be the entire thickness of the unsaturated zone, or the upper part of the profile affected by root abstraction or a fixed depth, e.g. 1 metre.

Field Capacity seems to be a straightforward, simple and practical concept and for many years has been accepted as such. However it has some very serious shortcomings.

In practice Field Capacity is a reasonable concept provided a. the soil is fairly conductive and b. it has a water table at a constant depth. In many cases however the depth of the water table varies during the year from, for example, several cm below the surface during a wet spell in the winter to a few m in the summer. Because the Moisture Content at Field Capacity varies according to the depth of the water table (see end of §2) the Field Capacity datum is not constant.

A further complication arises because the Hydraulic Conductivity can vary by several orders of magnitude over quite small charges in Moisture Content. Thus a stage can be reached during the drainage of the profile when the Moisture Content is such that the Hydraulic Conductivity begins to fall rapidly and further drainage is very slow (see equation 3.2). Drainage to the true equilibrium or Field Capacity may take a long time. A freely draining soil can reach true equilibrium with its water table in a day or two but a heavy clay can take several months by which time further rain will have fallen.

The more one considers the mechanisms controlling the flow of moisture in a soil the more one realizes that SMD and Field Capacity have very little physical meaning in most situations. Their merit is that they enable simple meteorological SMD Models to be constructed quite easily. The results from these models can be of value as a rough guide to the moisture status of the soil for such purposes as irrigation planning

A more scientific and physically-based approach to Soil Moisture Modelling is to work with the suctions and moisture contents rather than the SMD and to use an equation such as D'arcy's to describe the water flow from one point to another. It is essential to use such a model when dealing with moisture flow in the autumn when the ground is constantly draining and being re-wetted often with a water table of variable depth. Unfortunately these models are much harder to construct than SMD Models. They require information on the Soil Moisture Characteristic and the Hydraulic Conductivity vs Suction relation which are often not known accurately if at all. Also such models are more costly to run on the computer. They involve the solution of a number of equations on a discrete grid for a large number of timesteps.

Many of the ideas contained in this section are more fully described in references (1) and (2).

## 5. Field Drains

The purpose of field drains is to remove excess water from the soil so as to provide a. a more suitable environment for plant roots and b. drier conditions near the surface which enable machinery operations to be carried out more easily.

The depth of the drains and their spacing for a given soil type will often follow the tradition of the local area. Generally the drain depth is 70–85 cm in clayey soils and 1–1½ m in lighter soils. The spacing varies from 10 m in parts of the North to 40 m in some Eastern areas. Figures 5.1 and 5.2 describe the pattern of water flow in simple, idealized cases. It will be noted that the water table rises between the drains and that near the drain itself the flow is nearly radial into it with some water entering from below.

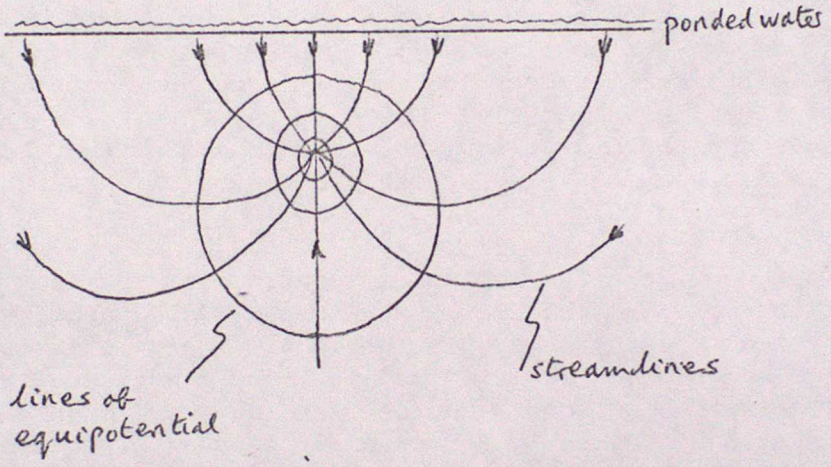


FIG 5.1 Steady-state flow pattern in an infinately deep homogeneous soil to a drain under a ponded surface.

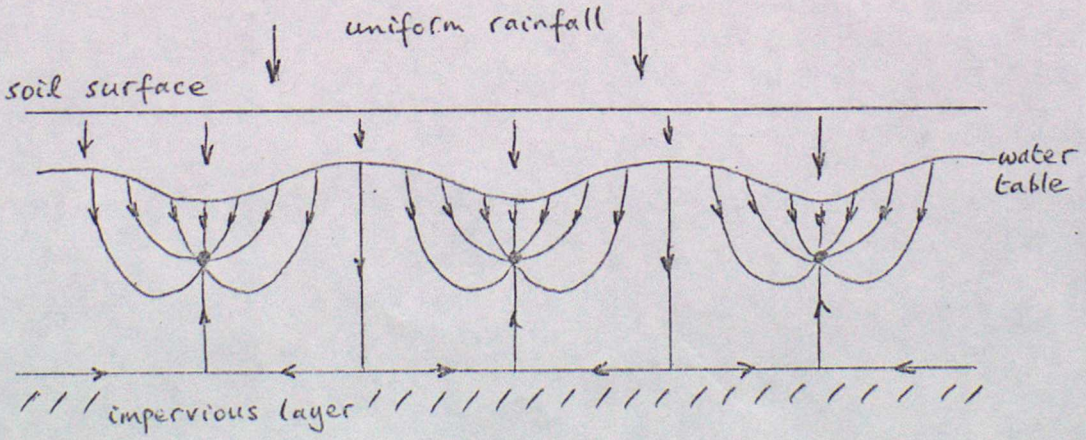


FIG 5.2 Steady-state flow pattern in a homogeneous soil where rainfall maintains the water table at constant depth. There is an impervious layer a finite distance below the drains.

Various drainage equations have been proposed which relate water table depth to rainfall input. Some of these, e.g. van Schilfgaarde's (3), require quite detailed information on soil properties and the geometry of the drain system which is often not available. One of the simplest and most widely applied equations is due to Hooghoudt (4). For a derivation of this equation see (6). Hooghoudt,s equation is based on the following assumptions,

i. the soil is homogeneous in all directions and therefore has a single, constant hydraulic conductivity function,

ii. the drains are parallel and equi-spaced,

iii. the suction gradient at each point beneath the water table is equal to the slope of the water table surface above that point,

iv. D'arcy's law applies,

v. an impervious layer exists below the level of the drains,

vi. there is a constant supply of rain and the system is in equilibrium.

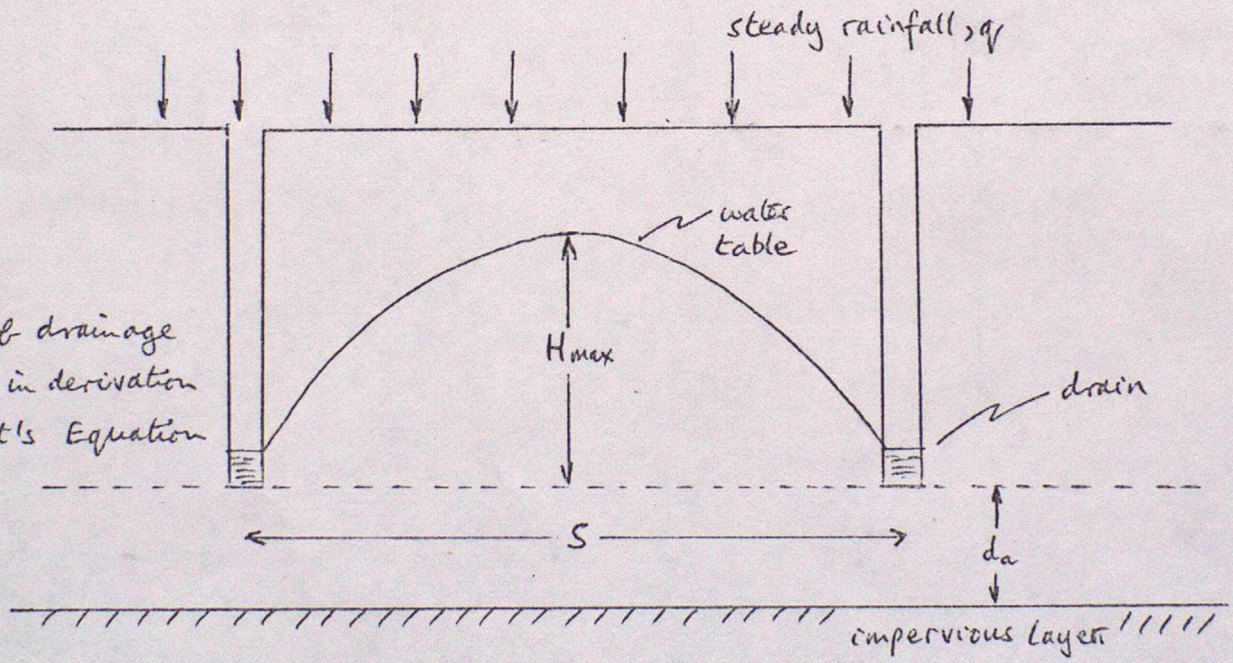


FIG 5.3

Geometry of drainage system used in derivation of Hooghoudt's Equation

Hooghoudt's equation is,

$$H_{max} = qS^2 / 8kd_a \qquad 5.1$$

Thus the height of the water table at mid-drain point is proportional to the rainfall rate, q for a given system.

Drainage design will often satisfy the criterion that "under a steady rainfall of 7-10 mm/day the water table will be maintained at a depth of 50 cm below the surface", ((5) and (11)). If the drain depth is 1 m and $H_{max} = 50$ cm with $q = 1$ cm/day we can simplify equation 5.1 further and write,

$$H_{max} = q/A \quad , \text{ where } \quad A = 0.02 \text{ day}^{-1} \qquad 5.2$$

This equation is sufficiently simple to provide the lower boundary condition to the soil-water system in a numerical model. Although not all of the assumptions will be strictly valid, particularly (i), a more sophisticated equation which is more specific would require more soil and drainage data than is normally available. Moreover a more complicated equation is difficult to accommodate into a numerical model.

In the case of clay soils which have a low value for $k$ the drain spacing needed to satisfy the above criterion is too close to be economically justified (approximately 1 m spacing is required). Mole drains are therefore introduced to feed water to the main drains below them, see figures 5.4 and 5.5.

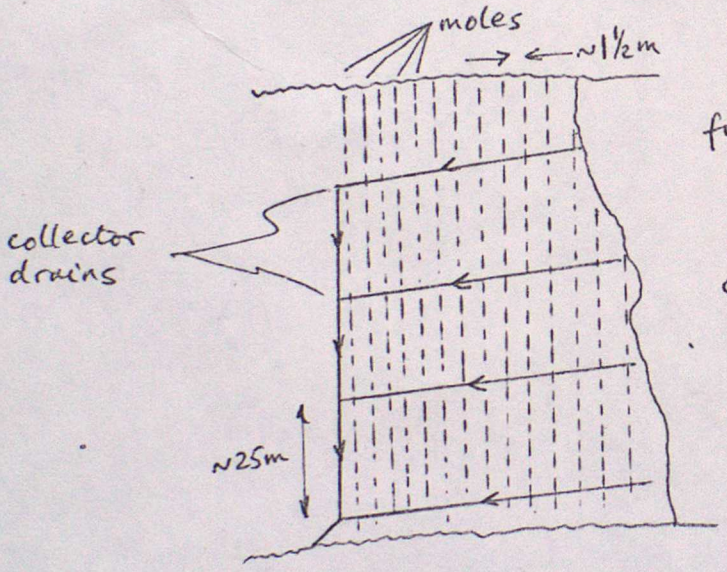


FIG 5.4 Plan view of Moles over Collector Drains

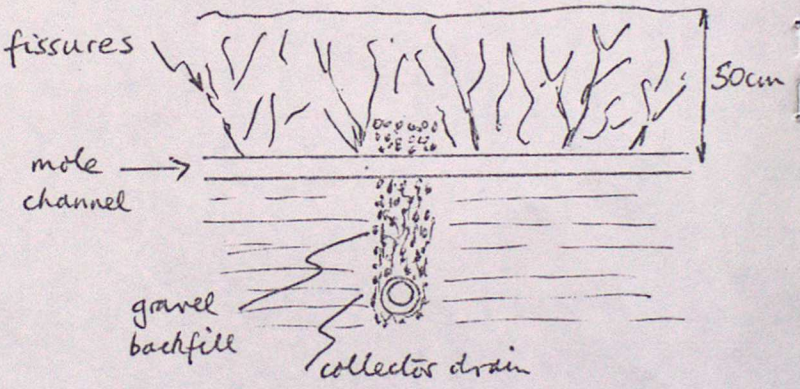


FIG 5.5 View along Collector Drain showing how Mole drain connects via gravel backfill

The geometry of the drainage system is now much more complicated than Hooghoudt's (Fig 5.3). However the system will approximately satisfy the criterion mentioned above. This enables the system to be modelled crudely using an equation such as 5.2 with an appropriate value for A.

## 6. Macropore Flow and 'Short-circuiting'

The simplest case of water infiltration into a soil to consider is the classical one-dimensional approach in which water moves downwards under the influence of suction gradients, these being no heterogeneity in the horizontal plane. D'arcy's law will then apply. However 100 years ago

Lawes et al. (7) found that a large part of the water added to soil profiles moved immediately through open channels and interacted only slightly with water in the soil itself. Unfortunately until recently this 'macropore' flow has been largely ignored but over the past few years much more attention has been paid to it in the literature (see (8) and (9)).

Soils with a high clay content will crack or fissure when dry. Heavy rain can flow down these fissures and into field drains by-passing the soil matrix. This has become known as 'short-circiting' (10). In modelling the flow of moisture, particularly in heavy soils, some attempt must be made to describe these processes, albeit simply.

References


1. Wellings S.R. and Bell J.P. 1982 Physical controls of water movement in the unsaturated zone. Q. J. Eng. Geol. London, 15, 235-241.

2. Bell J.P. and Wellings S.R. 1980? Soil Moisture Deficit, Field Capacity and all that. Internal publication from Institute of Hydrology, Wallingford, Oxon.

3. van Schilfgaarde J. 1965 Transient design of drainage systems. Drainage, Irrigation Div, Proc., Am. Soc. civ. Engrs, 91 (IR3), 9-22.

4. Hooghoudt S.B. 1937 Bijdregen tot de kennis van eenige natuurkundige grootheden van de grond, 6. Versl. Landb. Ond. 43, 461-676.

5. MAFF 1976 Climate and Drainage. Technical Bulletin 34.

6. Hillel D. 1980 Applications of Soil Physics Academic Press.

7. Lawes J.B., Gilbert J.H. and Warington R. 1882 On the amount and composition of the rain and drainage water collected at Rothamsted. William Clowes and Sons, Ltd. London.

8. Thomas G.W. and Phillips R.E. 1979 Consequences of Water Movement in Macropores. J. Environ. Qual., 8 (2) 149-153.

9. Beven K. and Germann P. 1981 Water flow in Soil Macropores II. A combined Flow Model. J. Soil Science 32 (1), 15-29.

10. Bouma J., Dekker L.W. and Muilwijk C.J. 1981 A Field Method for measuring short-circuiting in clay soils. J. Hydrol. 52, 347-354.

11. MAFF, ADAS 1979 Soil Examination and Management. (A booklet produced by members of Soil Science Dept. of ADAS, Shardlow).

12. MAFF 1975 Soil Physical Conditions and Crop Production. Technical Bulletin 29.

Any comments on this paper should be made directly to the author.

M Field
Meteorological Office, Met O 8a
London Road
Bracknell
Berkshire,
RG12 2SZ

15 September 1983